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A SURFACE WAVE DISPERSION STUDY OF THE CRUSTAL AND MANTLE STRUCTURE OF CHINA

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20. ABSTRACT (Continue on reverse side if necessary and identify by block number)

A measurement procedure was designed which can determine the group velocity with both good resolution and small systematic errors. This technique is being employed in a surface-wave study of the Eurasian continent. The generalized linear inversion theory is applied to obtain pure-path regional dispersion from mixed-path observations. Crust and upper mantle structure of the Eurasian continent is being interpreted with special emphasis on lateral heterogeneities.

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A SURFACE ANALYSIS ON MEASUREMENT AND INVERSION

As a continuing effort toward an improved surface wave dispersion study, our progress consists of the completion of the following two studies:

(a) A detailed study on the error and resolution of dispersion measurement, and

(b) a generalized inversion for laterally heterogeneous crustal and upper mantle structures.

MEASUREMENT

In carrying out our surface wave dispersion study, we are making use of the high-quality digital SRO data; a sample of it is shown by the top trace of Figure 1. This is the seismogram of an earthquake in Taiwan recorded at the SRO station GRFO in Germany. A typical surface wave train like this one propagating across a laterally heterogeneous crust and upper mantle results in a complex dispersive waveform that calls for special attention in numerically extracting its dispersion data.

The bottom two traces of Figure 1 are synthetic surface wave trains based on known complex wave spectrums. These two traces will be our calibration standards for various numerical methods discussed later.

Frequency-Time analysis (FTAN) is the widely used technique for dispersion measurement. Figure 2a gives the dispersion result of the top trace in Figure 1 using a constant relative bandwidth filter (CRBF) with a Gaussian parameter α=40 (Dziewonski et al., 1969). Figure 2b gives result using the optimum bandwidth filter (DBF) (Inston et al., 1971). In both figures, the solid dots are the maximum amplitudes which correspond to estimated group velocities. Also shown are two contours of 1 db and 10 db down from the maximum. These analyses demonstrated that using FTAN with a CRBF, which is most commonly used for dispersion measurement, might lead to poor resolution for certain period range.

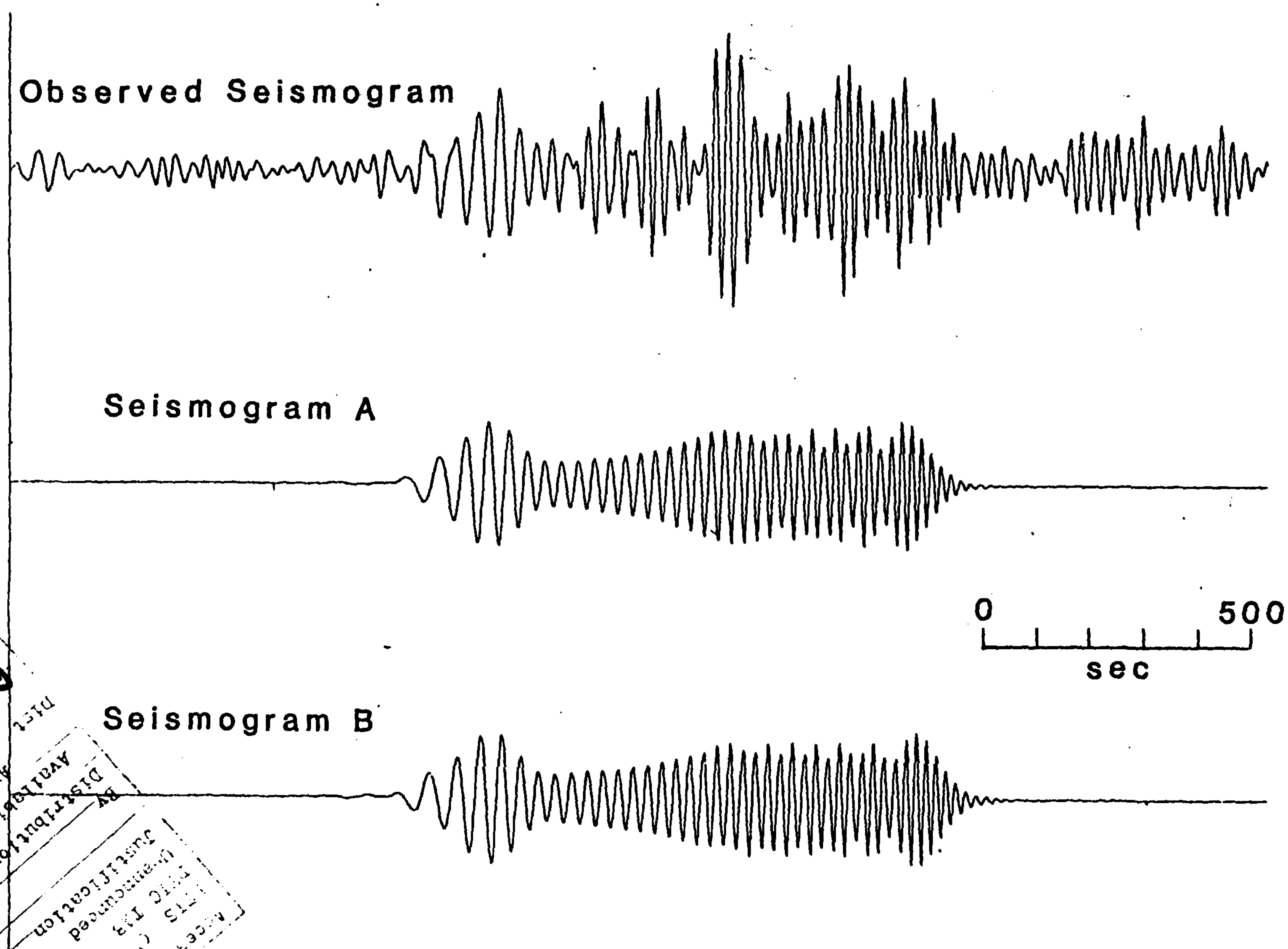


FIGURE 1

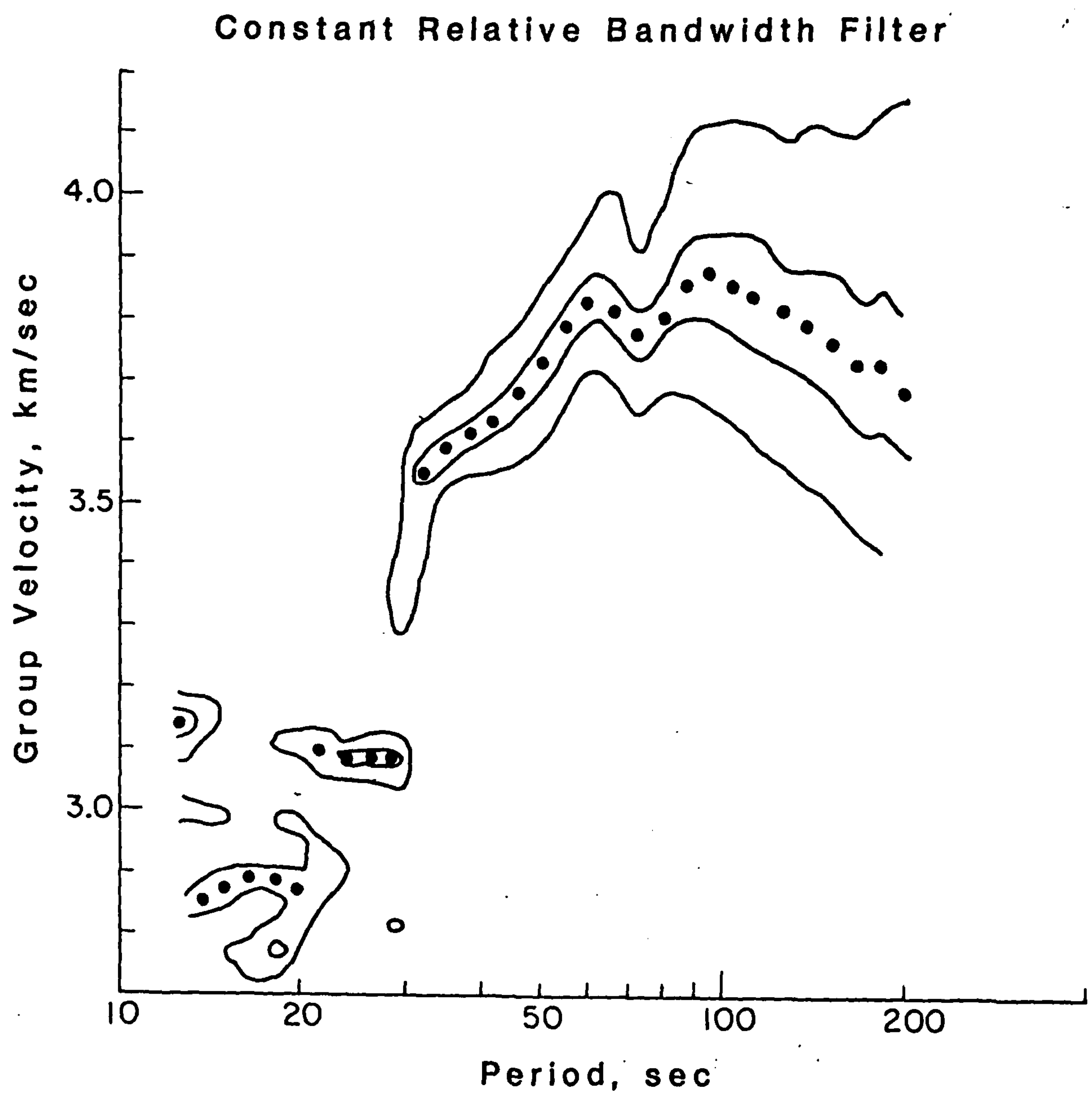
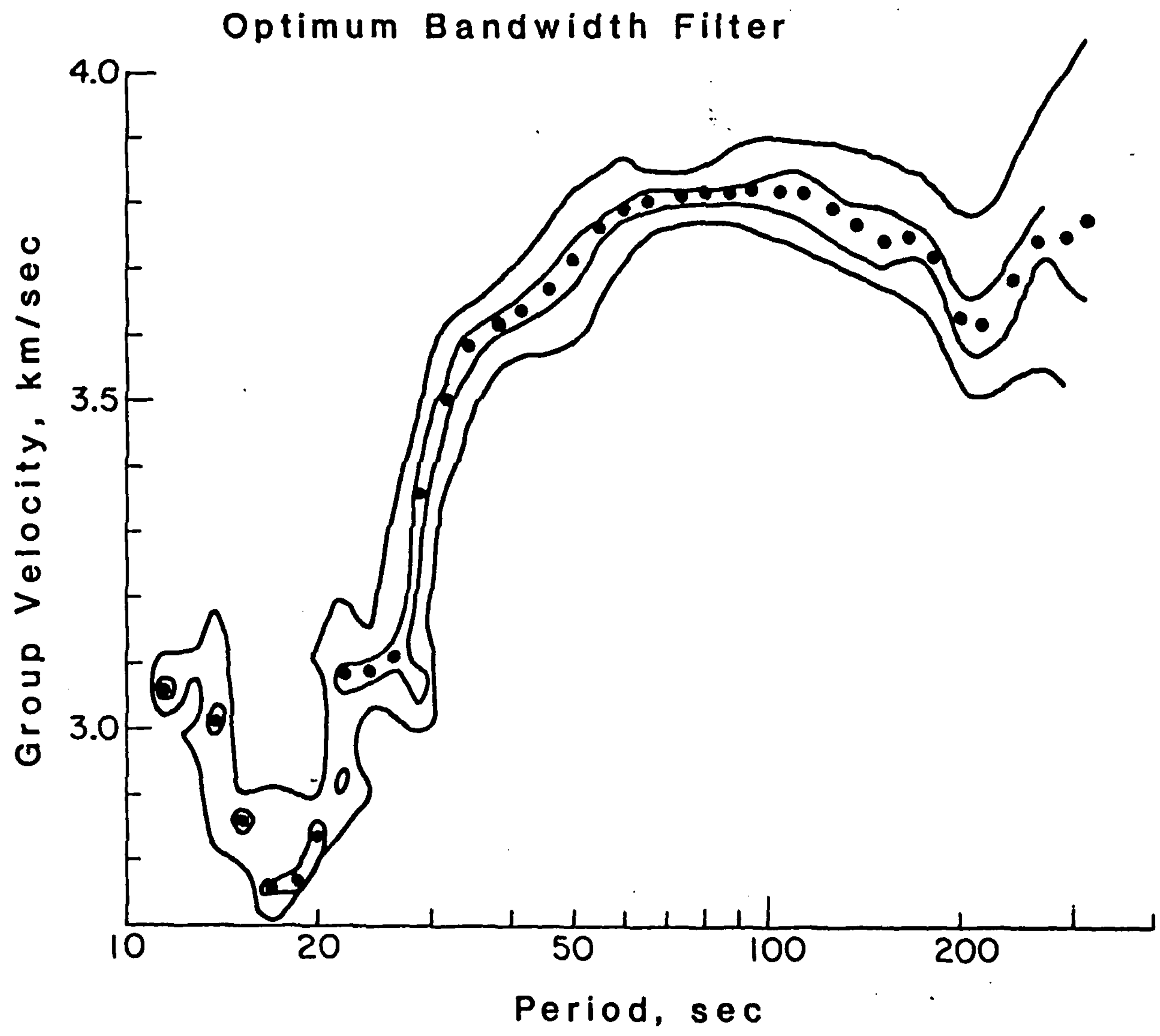

Constant Relative Bandwidth Filter
Group Velocity, km/sec
4.0
3.5
3.0
10
20
50
100
200
Period, sec

FIGURE 2a



Optimum Bandwidth Filter
4.0
3.5
3.0
Group Velocity, km/sec
10
20
50
100
200
Period, sec

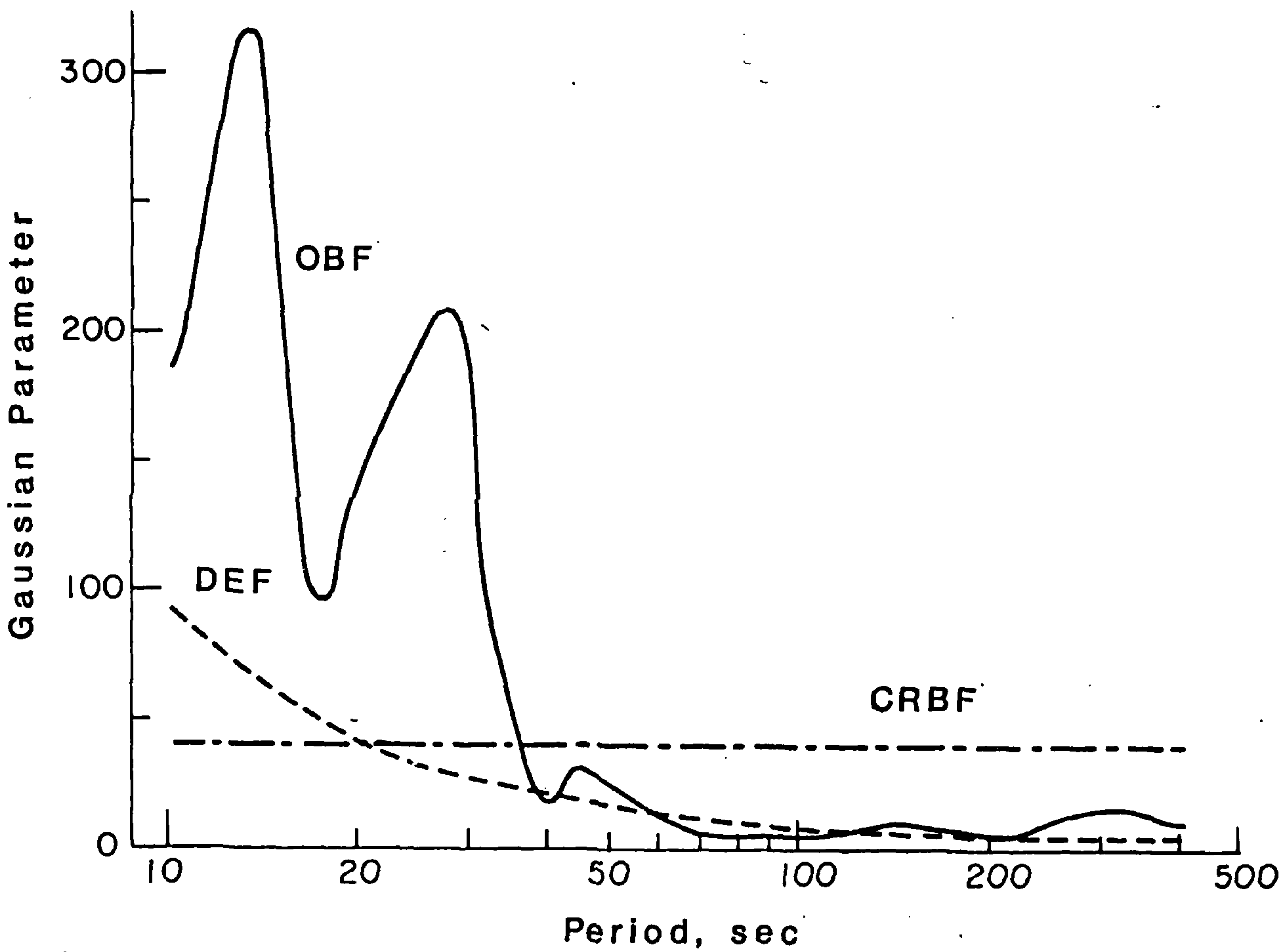
FIGURE 2b

For comparison, Figure 3 shows the Gaussian parameter α with respect to different filters as they are applied to this special seismogram. In addition to the curves of CRBF and OBF, a third curve shown here gives the result using the display-equalized filter (DEF) (Nyman and Landisman, 1977). The display-equalized filter is a Gaussian filter which has nearly equivalent bandwidth along velocity and log-period axes.

Then, the synthetic seismogram A is analyzed by FTAN to determine the systematic errors in group velocity. Systematic errors for CRBF and OBF are plotted against the period in Figure 4. For CRBF, we suspect that the significant distortion at short periods is due to poor resolution. The OBF is good for short periods. For long periods, however, a wider bandwidth is usually required to obtain good resolution which would introduce larger systematic errors. Because the dispersive property of the synthetic seismogram is smoother than that of actual seismograms, the systematic errors shown here are, therefore, smaller than those of the actual case.

At a given period, systematic errors become large if the amplitude and/or phase spectrums change rapidly at the neighboring periods. There are two ways to reduce systematic errors, one is to use larger α values, the other is to use the technique of match-filtering. The former is usually excluded because using a large α value generally results in losing the smoothing effect, and it is not uncommon to lose resolution completely. The technique of match-filtering is to measure the residual signal which is the cross-correlation of the observed seismogram with a theoretical signal. Because the residual signal is less dispersive compared to the observed seismogram, the determination of dispersion is more precise, with smaller systematic errors. This technique has been applied by Dziewonski et al. (1972) and is known as the residual dispersion measurement technique.

To take one step further than Dziewonski's work, a measured procedure



Gaussian Parameter
300
200
100
0
OBF
DEF
CRBF
10
20
50
100
200
500
Period, sec

FIGURE 3

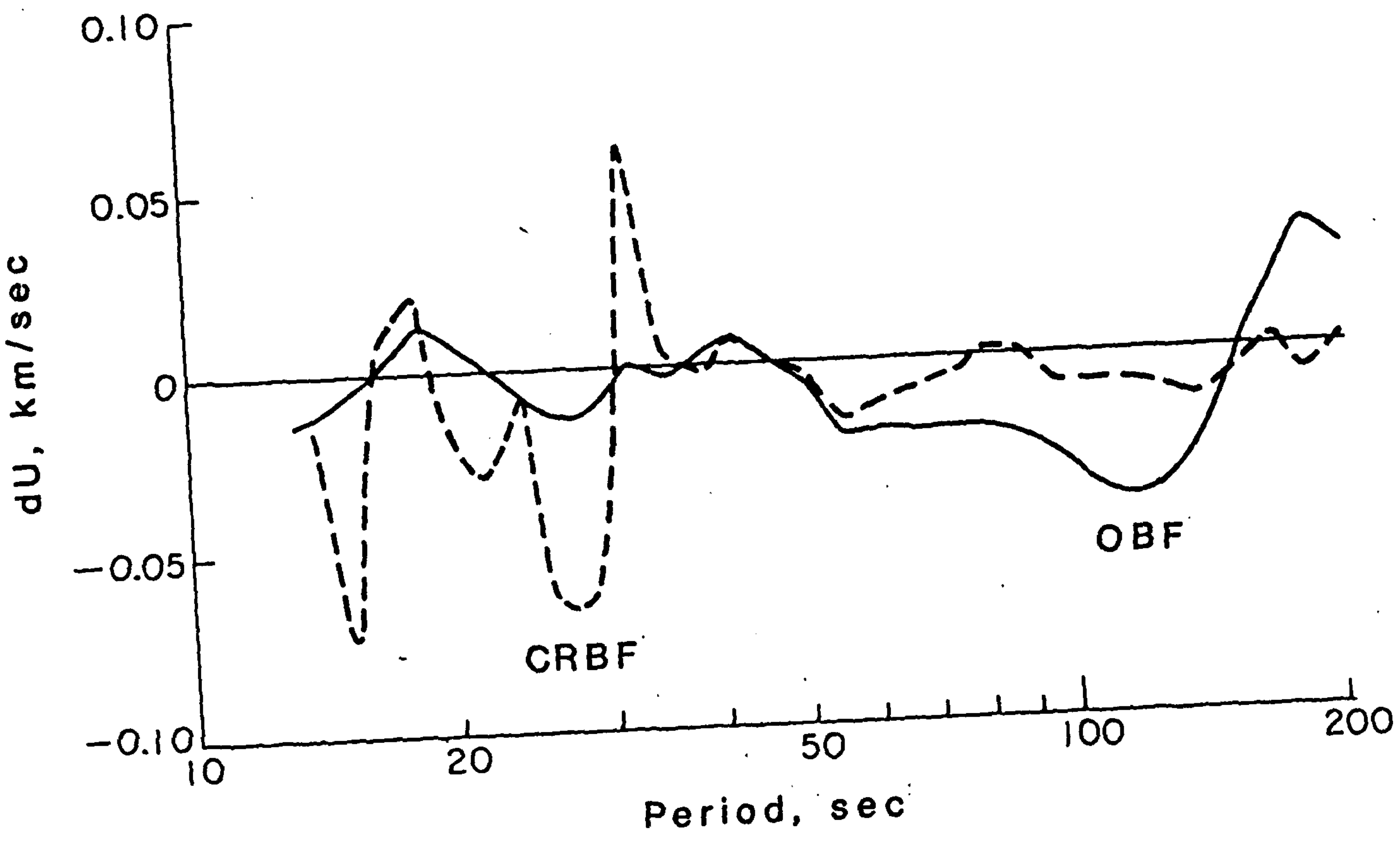


FIGURE 4

(Figure 5) is introduced which determines the group velocity with both better resolution and smaller systematic errors. We use the spectrum measured by FTAN with OBF as the theoretical seismogram because usually it does not have large discontinuities. Then the residual spectrum is constructed by taking the ratio of the observed spectrum to the theoretical spectrum. On top of the phase spectrum that Dziewonski treated, it is important to include the amplitude spectrum of the theoretical seismogram because the changing of amplitude is usually rapid at short periods. The residual spectrum is then analyzed by FTAN with DEF to determine the correction terms for the dispersion data. Because the residual spectrum is nearly white, a good resolution can be obtained by adding the correction terms to the theoretical dispersion.

This new measurement procedure has been tested using synthetic seismograms A and B. The systematic errors in group velocity have been reduced to about 0.01 km/sec (Figure 6).

Some examples of the application of this measurement procedure are given here. Figure 7 shows four travel paths from two earthquakes in Taiwan. These four SRO stations are SHIO, India, MAZO, Iran, ANTO, Turkey and GRFO, Germany. Among these four paths, only the path from Taiwan to Iran passes through the Tibetan plateau - a region of pronounced anomalous crustal and upper mantle structure.

Figure 8 shows the seismograms. From top to bottom are seismograms recorded at India, Iran, Turkey and Germany. The origin of the time scale has been set at the time corresponding to a velocity of 5 km/sec. The successive marks correspond to 4, 3.5, 3, and 2.5 km/sec.

Figure 9 shows the dispersion curves for these four paths; the x's are group velocities to India, crosses to Iran, triangles to Turkey and circles to Germany. Even if we exclude the path passing through the Tibetan Plateau, the dispersion curves for the other three paths are still quite different, inferring strong lateral heterogeneity. Therefore, it is inadequate to seek a single model for the entire Eurasian continent.

Measurement Procedure

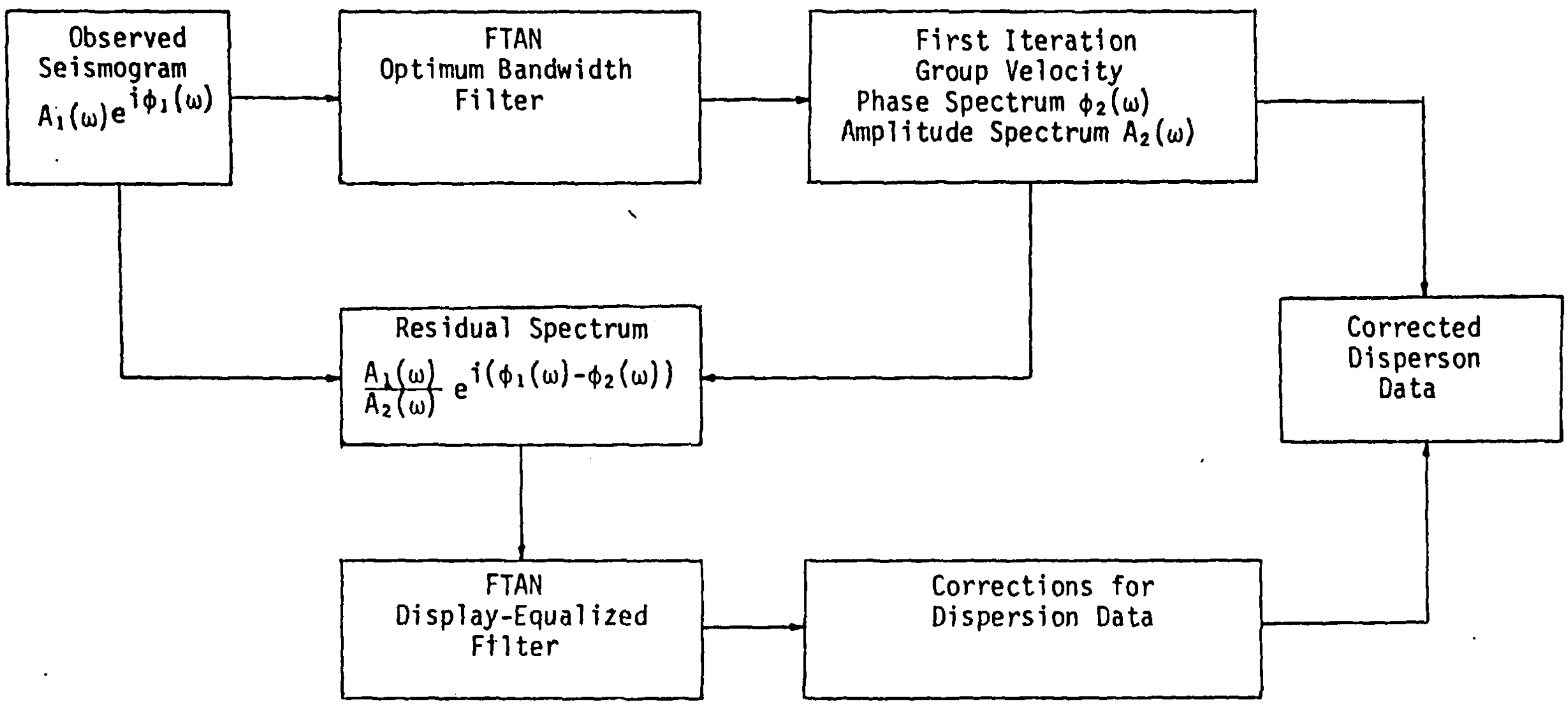


FIGURE 5

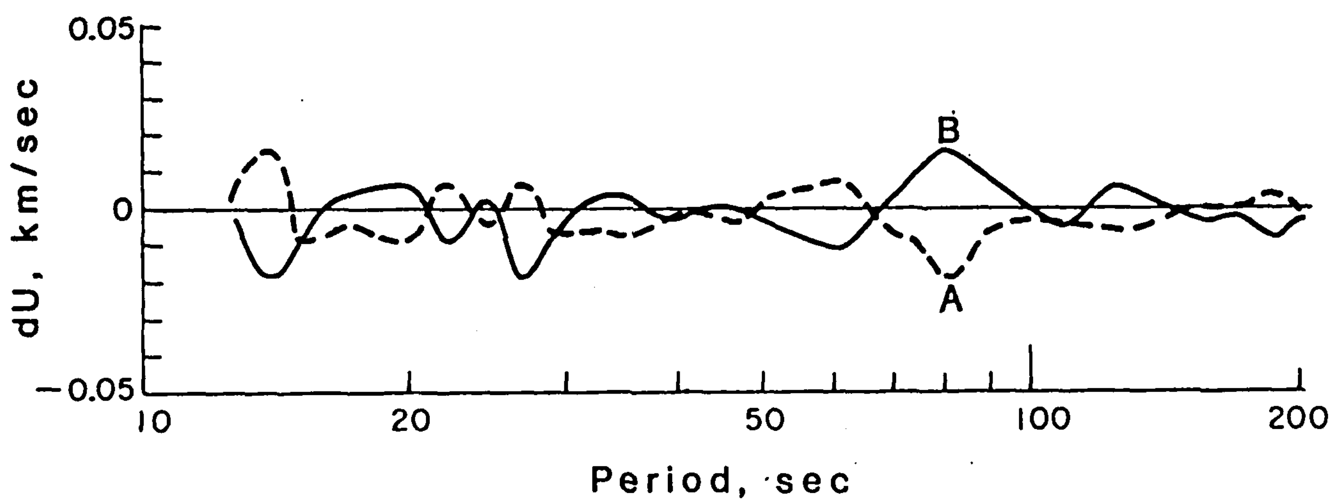


FIGURE 6

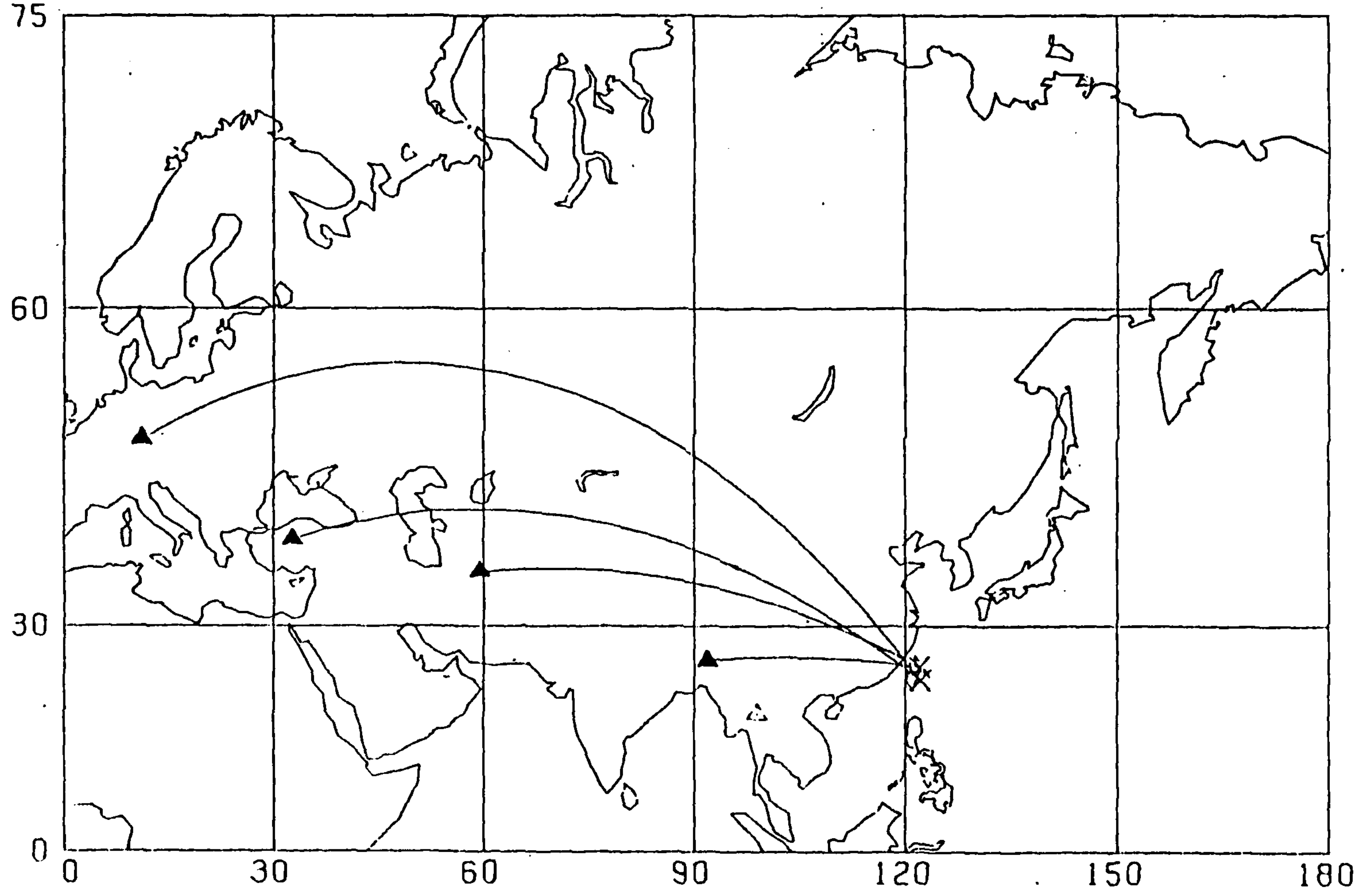


FIGURE 7

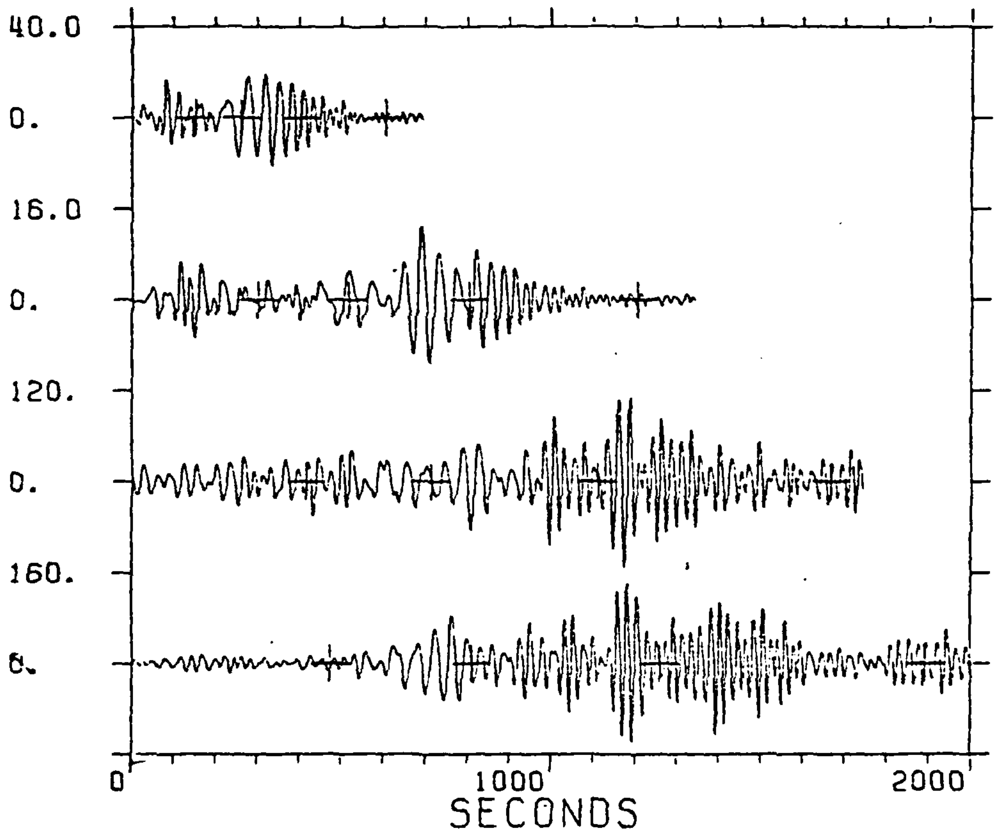


FIGURE 8

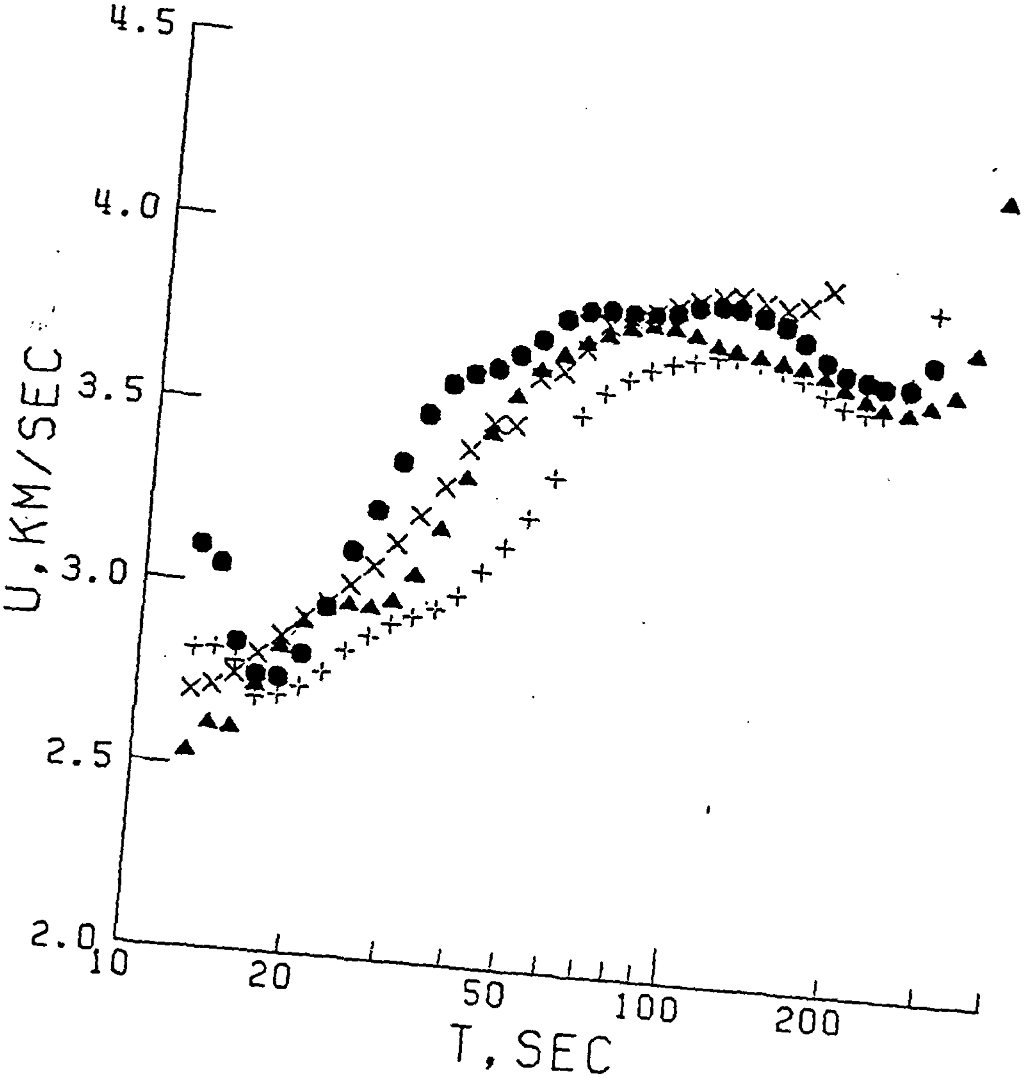


FIGURE 9

INVERSION

The generalized surface inversion technique now in common use is developed for horizontally layered halfspace. However, the dispersion data obtained in Eurasia are from a tectonic region which departs considerably from a horizontally layered halfspace model. Allowance must be given to various tectonic regions such as the Tibet plateau and the Mongolian shield massif. In general, surface wave paths crossing Eurasia typically traverse several tectonic provinces of distinctly different crustal, if not upper mantle, structures. These surface wave dispersion data thus indeed represent mixed-path results and an inversion based on a pure-path assumption can at best yield the average crustal and upper mantle structure over the entire surface wave path. The ideal of regionalization based on surface geological information subdivides a landmass into different tectonic regions; within each tectonic region the crustal and upper mantle structure is considered to be uniform (Wu, 1972; and Mills, 1978). To treat this subdivision problem in a more objective manner, a new technique is introduced which is called "grid-dispersion inversion".

With the "grid-dispersion inversion", the landmass is subdivided geometrically into geographical grids with grid size dependent on the dispersion information density. A typical starting grid network is shown in Figure 10 with 10° × 10° grid elements. This subdivision makes no use of surface geology information. It is hoped that upon the application of the "grid-dispersion inversion" both surface and subsurface geology will be recovered and the surface geology information would then serve as a redundancy check. The dispersion data appropriate for each grid are determined by the linear inversion theory in stochastic form (Franklin, 1970). After this linear inversion, the dispersion data for each grid are considered to be pure-path data which are individually used to invert for the crustal and upper mantle structure. The advantage of this technique is that as the number of observations increase, dispersion data for smaller grid size can be extracted, and a detailed study on lateral inhomogeneity can be performed.

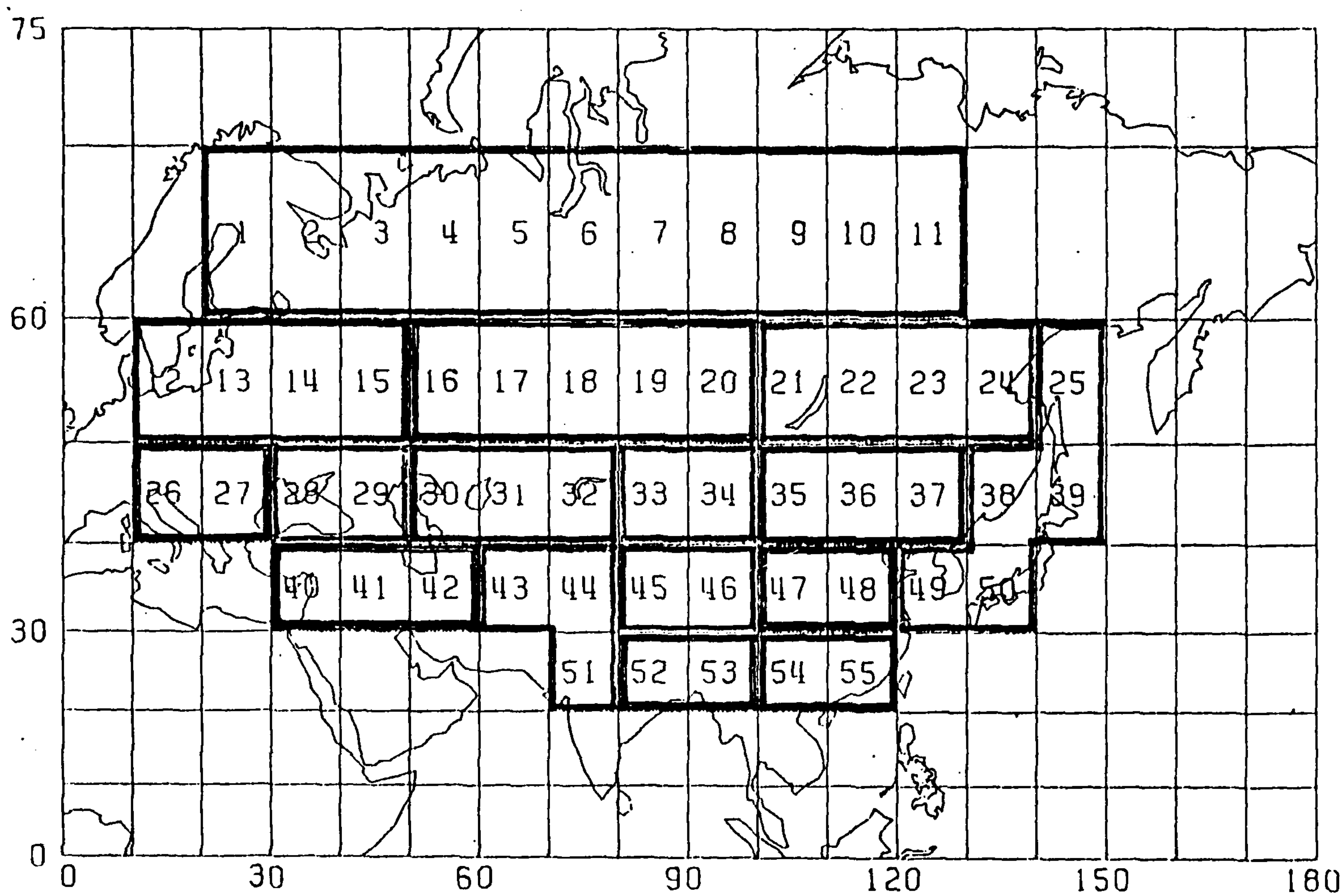

75
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FIGURE 10

To study lateral heterogeneities of Eurasian continent, a total of 54 paths have been measured (Figure 11). The Eurasian continent was initially subdivided into 55 blocks. An examination on the resolution matrix required that the number of blocks be degenerated to 16 (Figure 10). The pure path dispersion curves for these 16 blocks were determined. Two examples of the pure-path dispersion curves and the models derived will be given here: blocks 33 and 34 cover western Mongolia and Northwestern China; blocks 45 and 46 cover essentially the Tibetan plateau.

In Figure 12, circles are the pure-path group velocities of blocks 33 and 34 and triangles are those for blocks 45 and 46. These two data sets are inverted for shear velocity models, theoretical dispersion curves for these two regions are also shown in Figure 12.

In Figure 13, the dashed line is the model of blocks 33 and 34. No sedimentary layer is required to fit the data. The acidic layer and basic layer each have a thickness of 25 km, and both layers have the normal shear velocity values. The crustal thickness of this region is about 50 km.

The solid line in Figure 13 is the model of blocks 45 and 46. A 4 km sedimentary layer is favored by the data. The shear wave velocity of the acidic layer is lower than the normal value. The basic layer begins at a depth of 30 km and the crustal thickness reaches 70 km. The result is consistent with other published models, but our model is derived without any pre-set regional boundaries on surface geology.

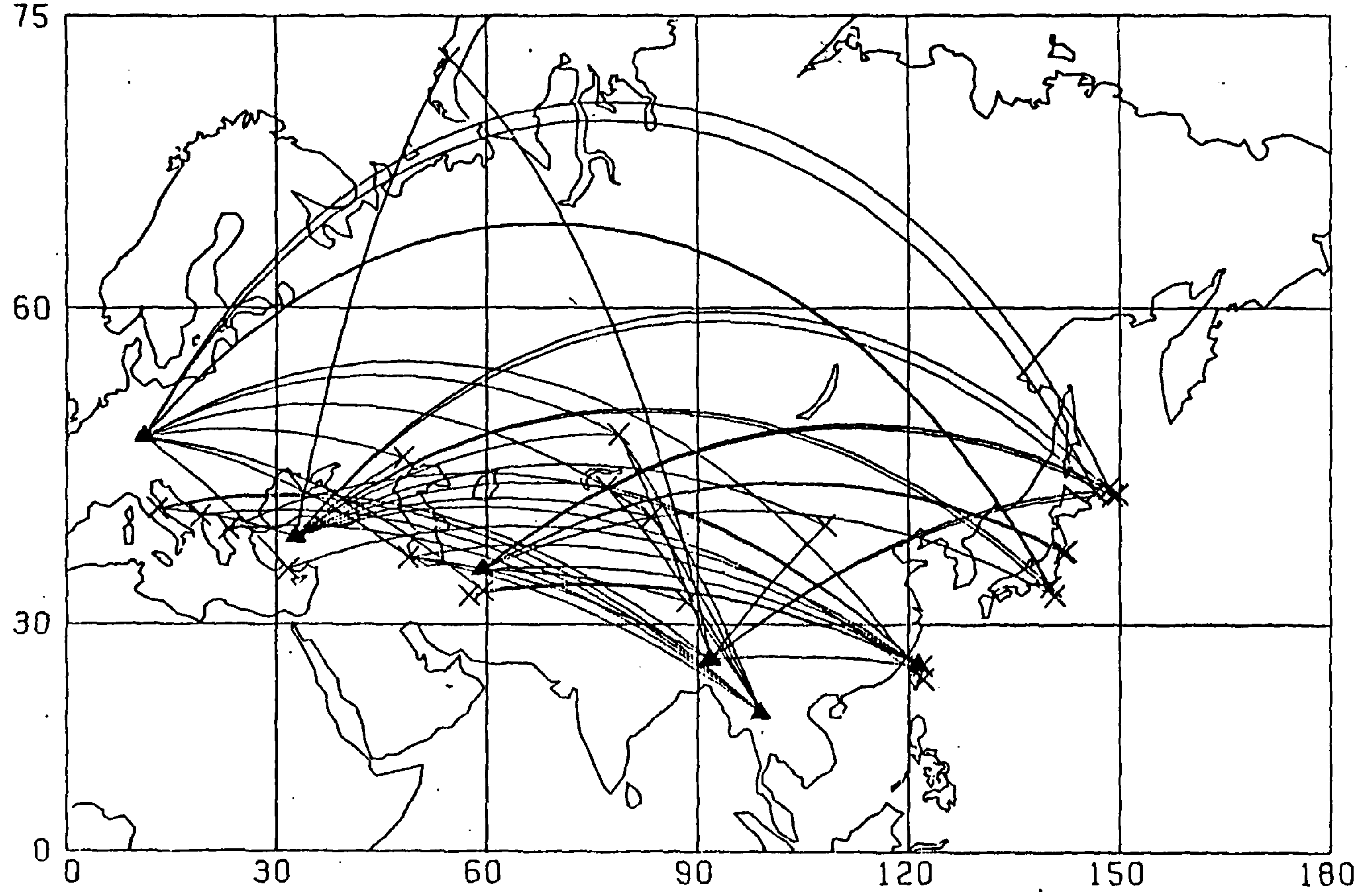

FIGURE 11

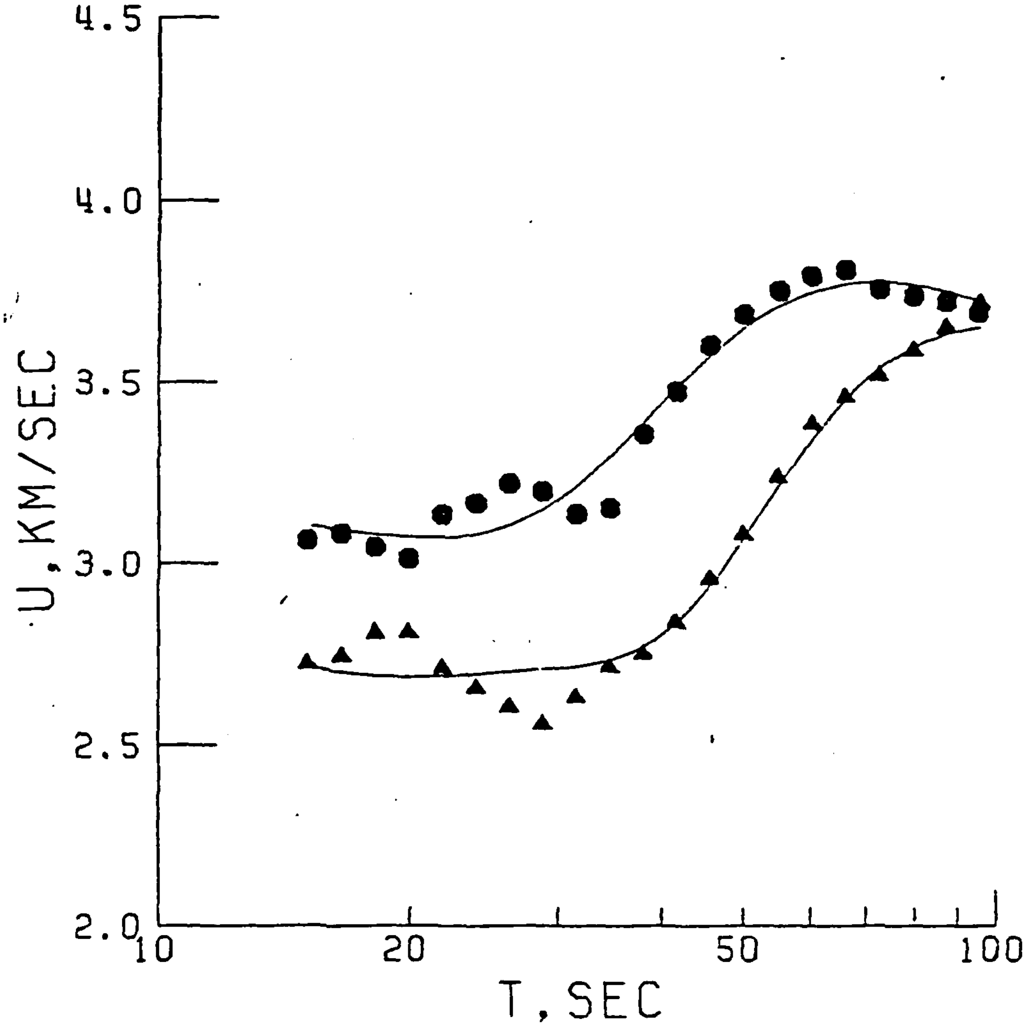

FIGURE 12

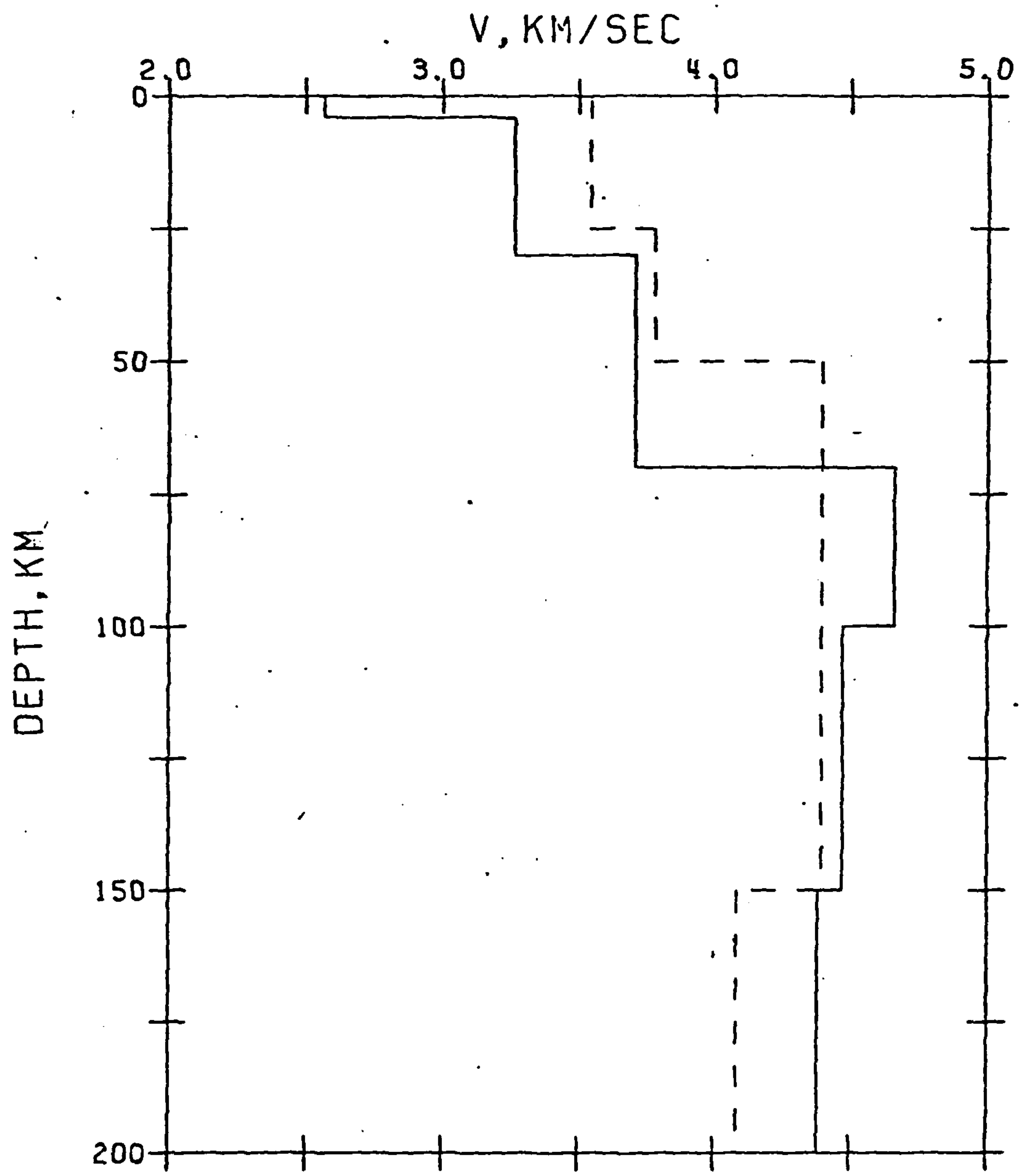

FIGURE 13

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